DOES OCEAN WATER REACH MANTLE PLUME SOURCES? IF SO WHAT ARE THE CONSEQUENCES?

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Plume basalts contain H2O and CO2, with possible sources: (1) the primordial Earth, (2) subducted lithosphere in deep mantle storage, or (3) the lithosphere. The depth at which volatile components dissolve in melt in a rising plume has consequences for trace element distributions.

The subduction of volatile components from the hydrosphere is a distinctive feature of convergent plate boundaries. These components are contained in the sediments and altered basalt and gabbro and serpentinites of the oceanic crust, and in hydrated peridotite of the subducted lithosphere or overlying mantle wedge. Large volumes of volatile components are removed by underplating of sediments at the accretionary prism, and by diagenetic and low grade metamorphic reactions. Rocks of blueschist or greenschist facies are carried deeper, and they experience a series of progressive metamorphic reactions, with facies inverted because of the thermal structure.

The broad picture was clear in the early 1970s: most of the H_2O stored in minerals of subducted oceanic lithosphere is lost through dehydration or melting reactions by a depth of about 100 km, or 150 km in favorable circumstances. What is known can be illustrated with old figures (Wyllie, 1973, 1979, 1984) and reference to recent treatments (e.g. Kesson and Ringwood, 1989; Peacock, 1990; Davies and Stevenson, 1990). **THE CRITICAL QUESTION IS WHETHER ANY HYDROUS MINERALS SURVIVE IN THE SUBDUCTED LITHOSPHERE DOWN TO DEPTHS WHERE DENSE, HIGH-PRESSURE HYDROUS MINERALS BECOME STABLE.**

The depths to which hydrous minerals can be carried is a function of their phase relationships (within their host rocks, not as isolated minerals), and the temperature variation as a function of depth and time within the subduction zone. Phase boundaries for the common hydrous minerals of the host rocks have been measured, and those for other candidates proposed as water carriers if not determined can be calculated. The thermal structure varies as a function of slab geometry, and time, with significant differences between a young subducted slab, and more mature slabs, and for a slab of young, warm lithosphere close to a ridge when subduction occurs. Temperature variations are represented in two ways: (1) The thermal structure at any time can be depicted by a set of isotherms, and variations with time are shown by a series of such diagrams. (2) Individual depth-temperature trajectories can be computed for specific parts of a subducting slab, and time variations can be shown by different trajectories, similar to the curves in Figs. 1 and 3. The positions of dehydration reactions are correspondingly illustrated by: (1) The dehydration fronts in mantle cross-sections, obtained by plotting the positions of dehydration reactions on the thermal structure maps as in figs. 2 and 4, and (2) The depths of dehydration of minerals located in particular parts of subducted lithosphere, given by the intersections of depth-temperature trajectories with dehydration boundaries, as in Figs. 1 and 3.

Many thermal structures have been calculated for subduction zones. Two early versions covering a wide range of possible conditions provide the dehydration fronts shown in Fig. 2. Fig. 3 shows three bands for the temperature variation within a 5 km-thick ocean crust, OT and TMJ, corresponding to curves in Fig. 1, and the third band, ADS, corresponding to a calculation in which the crust is strongly cooled by endothermic dehydration reactions, even cooler than Peacock's (1990) coolest trajectory for 200 my old lithosphere, which is near values of Davies and Stevenson (1990). Very high thermal gradients exist in the subducted oceanic crust, with cooler rock situated below warmer rock.

Figures 2 and 4 show the positions of dehydration fronts where H_2O is released from the most common hydrous minerals expected in the three subducted materials. Fig. 2 shows the maximum depths to which the hydrous minerals could be carried if H_2O was available, and Fig. 4 shows the orientation of the dehydration fronts in the subducted crust. The positions of the reaction boundaries vary as the thermal structure changes, and the serpentinite reaction (DG) (approximately equivalent to extrapolation of the greenschist facies boundary) may be stretched out through a considerable depth interval, or compressed into a much shorter interval (Wyllie, 1979, 1984). Amphibolite breakdown, on the other hand, is a pressure-sensitive reaction under these conditions (Figs. 1 and 3), and its position within the oceanic crust can change

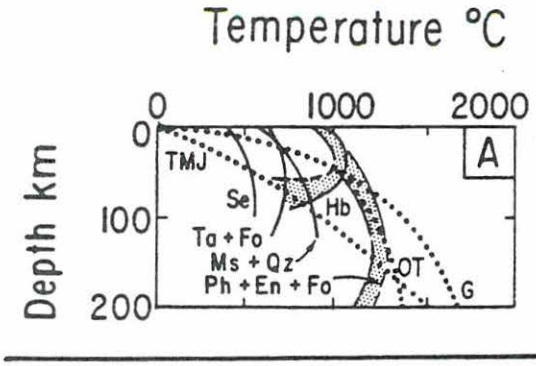

FIG. 1. Dehydration reactions, and estimated depth-temperature trajectories (TMJ, OT, G) along upper boundary of subducted lithosphere slab. Compare with Fig. 3. Se = serpentine, Ta = talc, Fo = forsterite, Ms = muscovite, Qz = quartz, Ph = phlogopite, En = enstatite, Hb = amphibole (Wyllie, 1973, Fig. 3A).

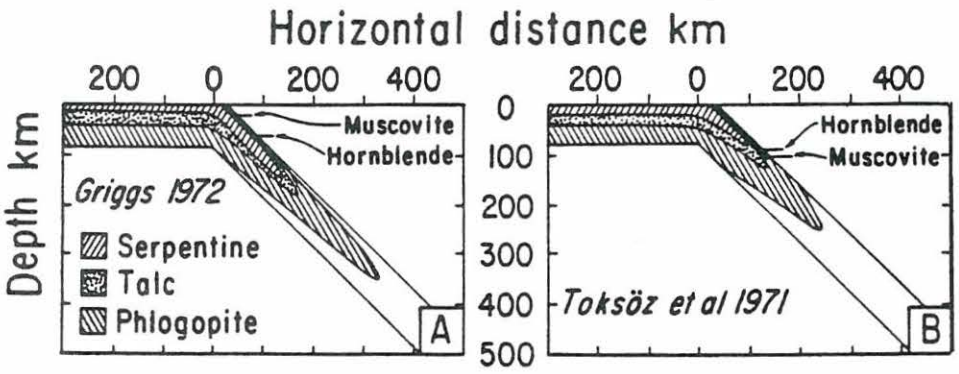

FIG. 2. Schematic cross-sections through subducted slab, showing maximum depths to which hydrous minerals can be carried, if the host-rock composition is appropriate (with H_2O), according to two thermal structures (see G and TMJ in Fig. 1 for slab-surface trajectories). For a view of the dehydration fronts (metamorphic facies boundaries) within the oceanic crust, see Fig. 4. (Wyllie, 1973 Fig. 7).

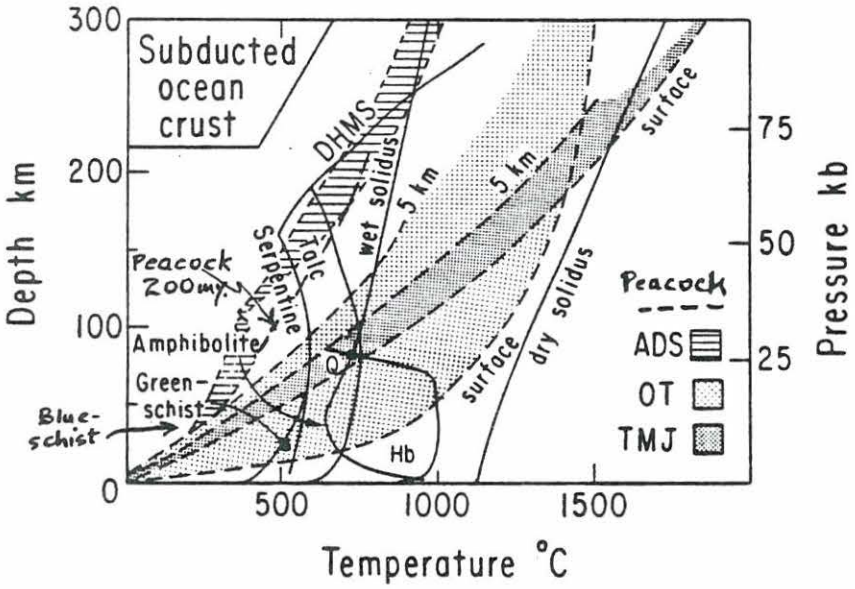

FIG. 3. Phase boundaries for subducted oceanic crust of gabbro, enclosing bodies of serpentinite. The shaded bands give temperature distribution across a 5-km thick subducted crust according to three investigations. Compare OT and TMJ with Figs. 1 and 2. ADS trajectories are even cooler than the coolest calculated by Peacock (1990). Selected dehydration reactions as in Fig. 1, plus dense hydrous magnesian silicates (DHMS), compared with solidus curves dry and wet. Note the blueschist, greenschist, and amphibolite facies. Compare dehydration fronts and facies in Fig. 4 (Wyllie, 1979, Fig. 15).

through only a limited range.

Pelagic clays yield micas when metamorphosed, and breakdown or melting of muscovite + quartz may represent the conditions for almost complete dehydration of pelagic sediments, at depth less than 100 km (Fig. 2). For the coolest slabs (Peacock, and ADS in Fig. 3), mica could survive to somewhat greater depths.

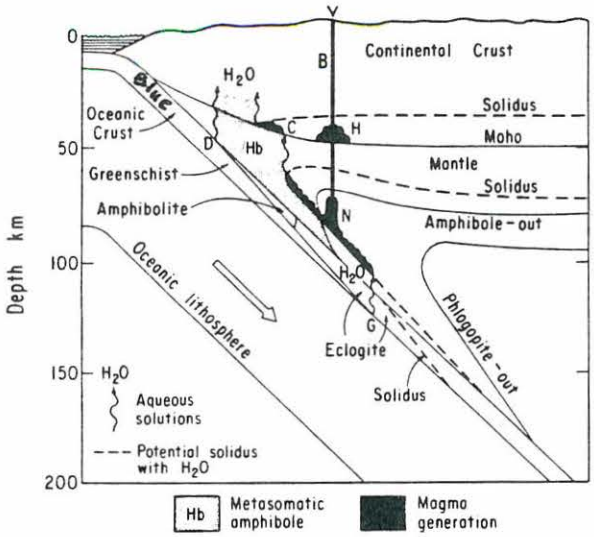

FIG. 4. Selected reaction boundaries plotted on subduction zone thermal structure, for cool subducted crust and warm mantle wedge, based on dehydration reactions in Figs. 1 and 3. Compare Fig. 2. Note the greenschist dehydration front, DG, corresponding also to serpentine dehydration. The amphibolite dehydration front is the short line between amphibolite and eclogite facies. Melting begins if H_2O migrates across the dashed solidus curves. Reaction positions change with thermal structure (Wyllie, 1984, Fig. 5).

Amphibolite dehydration or partial melting at depth near 80 km yields anhydrous eclogite. Serpentinite bodies in the bottom of the oceanic crust may carry water to depth G in Fig. 4. The geometry of the serpentinite boundary, DG (Fig. 4), is sensitive to the thermal structure, but it appears to be difficult to force the position of point G any deeper than 180 km (see the coolest trajectory, ADS, in Fig. 3). If the subcrustal oceanic lithosphere does become hydrated by deep penetration of ocean near the ridges or trenches, serpentine and talc appear to be limited to shallower than 180 km even in the centers of the coolest lithosphere slabs (Fig. 2A). Phlogopite is the only common hydrous mineral with a greater depth range. As shown in Figs. 1 and 4, phlogopite in oceanic crust or overlying mantle (formed possibly by reaction with solutions or melts from the subducted slab; Wyllie, 1984, Fig. 6) could reach a maximum depth of about 180 km, where it would dehydrate or melt. The deeper field for potential phlogopite within the center of the subducted slab is unlikely to be realized in depleted, low-K harzburgite.

THERE HAS BEEN SPECULATION THAT OTHER HYDROUS MINERALS FORMED AT LOW TEMPERATURES MIGHT SURVIVE TO GREATER DEPTHS IN: (1) BLUESCHIST FACIES PELAGIC CLAYS, AND (2) BLUESCHIST FACIES GABBROIC CRUST (PEACOCK, 1990), AND (3) SUBCRUSTAL SERPENTINIZED HARZBURGITE (KESSON AND RINGWOOD, 1989). THERE IS A REAL PROSPECT THAT CARBONATE MAY ESCAPE DISSOCIATION OR MELTING FOR LONG-TERM RESIDENCE IN THE MANTLE (HUANG ET AL., 1980).

Peacock (1990) listed the following candidates for pelagic clays and gabbros: phengite, glaucophane, chlorite, lawsonite, stilnomelane, clinozoisite, epidote, laumontite, lawsonite, margarite, prehnite, waikarite, and zoisite. Stern and Wyllie (1973) studied a pelagic red clay at 30 kbar and found only one subsolidus hydrous mineral at 675°C, a phengitic mica. However, the occurrence of other blueschist minerals at lower temperatures along the trajectories ADS and Peacock (Fig. 3) remains a possibility. Schreyer et al. (1991) considered newly sythesized high pressure hydrous minerals such as MgMgAl-pumpellyite. Candidate minerals proposed have not yet been demonstrated to be stable in suitable bulk compositions to appropriate pressures and temperatures, but the prospect exists.

There is a family of dense hydrous magnesian silicates stable only at high pressures (DHMS in Fig. 3), which include: Phase A, a 10 angstrom phyllosilicate, and clinohumite. If hydrous minerals being subducted survive until they enter this region then, when they dehydrate, some water can be recaptured in DHMS for transportation to greater depths. Kesson and Ringwood (1989) appealed to dehydration of serpentinite through the depth interval of 150-300 km, proposing that subcrustal serpentinite is carried deep enough to enter the DHMS field, with more dehydration occurring to 400 km. Fig. 4 shows that even with the coolest oceanic crust (ADS), serpentine and talc are completely dehydrated between about 160 km (top) and 180 km (bottom) before the dense hydrous magnesian silicates (DHMS) become stable. However, the P-T trajectory for the deepest crust is very close to reaching the DHMS field as dehydration is

completed, and if serpentine is present within lithosphere beneath the oceanic crust, this cooler lithosphere could follow a trajectory passing directly from serpentinite into the stability field for DHMS, where dehydration of the serpentinite and talc would leave some bound water in DHMS.

THE EVIDENCE SUPPORTS A CONCLUSION THAT NORMALLY THE SUBDUCTED OCEANIC LITHOSPHERE IS COMPLETELY DEHYDRATED BY A DEPTH OF 150 KM. THE ONLY PROSPECTS FOR DEEPER SUBDUCTION OF SMALL QUANTITIES OF OCEAN WATER APPEAR TO BE IN THE OLDEST, COOLEST LITHOSPHERE, IF: (1) SOME HYDROUS MINERALS FROM THE BLUESCHIST FACIES ESCAPE DEHYDRATION OR MELTING, OR (2) EXCEPTIONALLY DEEP HYDROTHERMAL ALTERATION OF SUBOCEANIC MANTLE CARRIES H_2O THROUGH SERPENTINE AND TALC FIELDS TO ONE OF THE DHMS MINERALS.

The mantle wedge is a site where the lithosphere is thoroughly impregnated by solutions and, under some conditions, by hydrous siliceous melts (Fig. 4).

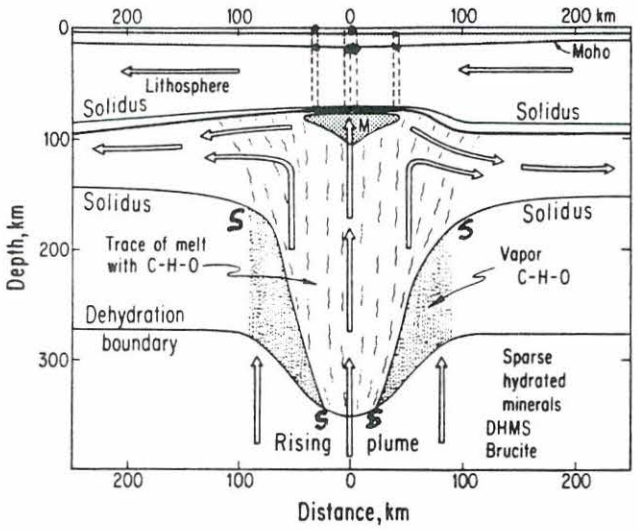

FIG. 5. Distribution of vapors and melts in mantle plume obtained by superimposing extrapolated phase boundaries for peridotite-H_2O-CO_2 on isotherms in plume thermal structure. The locations of phase boundaries vary with temperature and oxygen fugacity, but the general structure persists. In a rising plume, volatile components are released from solids at depths below about 300 km. The cylindrical plume contains a core with a trace of interstitial volatile-rich melt, and a sheath of vapor. There is a small kernel at the top of the plume, M, where significant melting occurs, forming picrite. See Wyllie (1988) for assumptions and construction.

Consider now the reversal of subduction, with a mantle plume carrying volatile components from the interior toward the surface. The DHMS boundary in Fig. 3 confirms that water if present deeper than 300 km would be stored in minerals, as indicated in Fig. 5 (if oxygen fugacity is low enough, a vapor with CH_4 or even H_2 may be present). Volatile components are released as the plume crosses the DHMS boundary (follow P-T trajectories from high to low pressure in fig. 4), and the vapor rises with the plume until it crosses the solidus, where it becomes dissolved in a trace of melt. The melt is rich in volatile components and incompatible elements. The line around the shaded area M corresponds to the solidus for dry peridotite, and significant melting with generation of picritic melt occurs only in this limited volume. The distribution of trace elements and isotopes within and above the plume are influenced by the different partition coefficients between minerals and (a) vapor, (b) volatile-rich melt, and (c) picrite. Trace element concentrations in volatile-rich melt are set at different pressures as a function of the slope of the solidus boundary (S-S), i.e. as a function of distance from the center of the plume. These differences would persist in the melt reaching the lithosphere at different distances from the plume center.

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